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SCENES IN OHIO DURING THE LAST ICE AGE†

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During the last decade more than a dozen geologists and some pedologists and botanists have tackled the surficial deposits of western Ohio for a reinterpretation of Ice Age history. This was first deciphered in detail at the turn of the century by Leverett (1902). The new ideas are embodied in the new Glacial Map of United States East of the Rocky Mountains (Geological Society of America, 1959) and a forthcoming Glacial Map of Ohio (U. S. Geological Survey). During this same decade the young science of glaciology has mushroomed due to military studies of polar ice caps and to the International Geophysical Year. It seems entirely appropriate to put these together; in short, what can today's ice caps tell us of Ice Age events? The most complete record in surficial deposits is that of the very last Ice Age, the classical Wisconsin Stage. What was this last North American ice cap like?

This application of glaciology to Wisconsin Stage history follows the principle of uniformitarianism. It is held that whatever took place in Ohio 9,000 to 30,000 yr ago is taking place somewhere in the world today. Several specific assumptions are involved: (1) Major weather conditions in Wisconsin time over the ice sheet in North America were essentially like those today over the much smaller ice caps at higher latitude. Climatic zones were squeezed together, thus intensifying winds, over mid-latitudes like Ohio, but the same storms we have today must have delivered snow to the former ice cap. (2) Glacial ice flowed in essentially the same fashion 20,000 yr ago as it does today. This is implicit in the similar stress values at the bottom of a great variety of arctic and temperate glaciers as calculated recently (Goldthwait, 1957). (3) Topographies similar to Ohio can be found under Malaspina Glacier in Alaska, in some broad valley areas of Glacier Bay; so, these affect ice today much as Ohio topography affected Wisconsin ice. (4) Radiocarbon measurements of time, so numerous in Ohio, are essentially correct at least as to relative order. They are internally consistant in Ohio, albeit different from the preconceived conventional pattern.

Advance of the Last Ice into Ohio

Conditions near the ice edge.—In many spots near the outer part of an ice sheet, the glacier covered but did not destroy the prior land surface; so, there is some record of the terrain into which the ice moved. The deposits near Cleveland show it was a lake dammed before advancing ice (White, 1953); carbonaceous alluvium at Hamilton suggests it was a flood plain; rotted till at Brush Creek suggests it was swampy (Forsyth, in preparation); and at several other exposures there is very thin buried forest soil (Goldthwait, 1958). Each of these surfaces is on earlier glacial drift. The surface was and is gently undulating or flat just as would be the case if ice were to spread over Ohio today; the deep preglacial bedrock valleys were already filled by glacial deposits.

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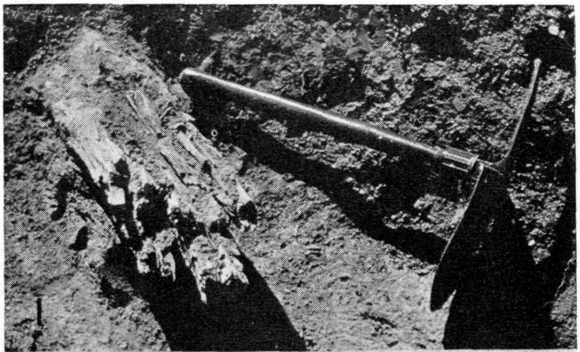
The environment was different as regards vegetation and climate, however. Laid out on the paleosol at Sidney (LaRocque and Forsyth, 1957) and in the lake at Cleveland are spruce logs (fig. 1). The forest litter at Oxford contains spruce needles, cones, and fly wings. At 15 places now similarly-dated spruce wood has been found within the upper glacial till. These involve hundreds of logs,



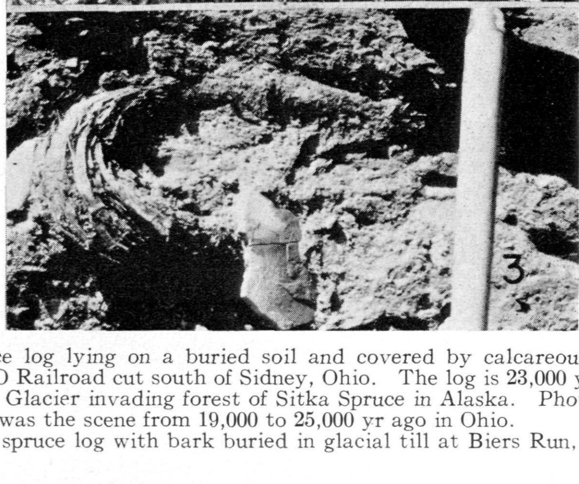


FIGURE 1. Spruce log lying on a buried soil and covered by calcareous blue-gray till at B & O Railroad cut south of Sidney, Ohio. The log is 23,000 yr old.

FIGURE 2. Taku Glacier invading forest of Sitka Spruce in Alaska. Photo by W. O. Field, This was the scene from 19,000 to 25,000 yr ago in Ohio.

FIGURE 3. Bent spruce log with bark buried in glacial till at Biers Run, Chillicothe, Ohio.

sometimes packed together in great profusion (fig. 2). These trees evidently were alive when ice mowed them down because some were bent double without breaking, and many still retain tight bark (fig. 3). Old dead logs snap when bent. The spotty grouping of many logs suggests that these trees were in patches or perhaps prolific along streams. As Martin (1959) has recently deduced, the tree size suggests a taiga or open forest type of association, such as is now found at Knob Lake near 54° North Lat. in Quebec (fig. 4). If identical, the mean annual temperature then was 23°F which is 30° cooler than is Columbus today (table 1). More likely the warmest month was like July at Knob Lake, mean daily temperature near 55°F. It may be fair to surmise that, as the ice approached, July temperatures in central Ohio were about 20° cooler than today.

Precipitation probably was heavier because the glacier had to be well-fed with snow. Nevertheless, in comparable forest environment at Knob Lake, mean annual precipitation is only 32 in.; but Juneau surrounded by active glaciers gets 83 in. annually. The large ice masses at lowest latitudes today receive 60 to 110 in. on their upper surfaces (Ahlmann, 1939; Diamond, 1958; Sharp, 1951). Although Ahlmann's original studies (1940) blamed temperature mostly for glacier fluctuations, more and more glacial budget studies place the emphasis of glacier size on the amount of actual snowfall rather than on low temperatures per se.

TABLE 1

Today's climate where the glacier used to be

	Mean daily temperature			Mean annual precipitation	
	Annual	January	July	All kinds	Snowfall
Former ice source areas in Canada					
Knob Lake Area	23°F	−11°F	55°F	32 in.	121 in.
L. Mistassini Area	31°	− 6°	60°	35 in.	122 in.
North Bay Area	38°	10°	67°	33 in.	81 in.
London, Ontario Area	47°	22°	70°	35 in.	40 in.
Former glaciated areas in Ohio					
Cleveland, Ohio Area	51°	28°	73°	32 in.	36 in.
Columbus, Ohio Area	53°	30°	75°	37 in.	16 in.
Hamilton, Ohio Area	54°	32°	76°	40 in.	15 in.

How the ice spread.—Wherever the last ice went, it spread the ubiquitous glacial *till* which is found in back yards in 63 counties in Ohio (fig. 5). To the relatively smooth earlier glacial deposits, it simply added a veneer of till one to 90 ft thick (fine rock-flour, silt, sand, pebbles and subangular stones all mixed). There is reason to believe that this till was smeared down from beneath moving ice (fig. 6) because laborious "till fabric" measurements in a dozen places in Ohio show that the long axes of most stones trend in one preferred direction (Holmes, 1941). Perhaps stones were oriented by shearing in basal ice which melted from below as it stopped flowing (Harrison, 1957). In cold arctic ice, studied in tunnels in northwest Greenland, the glacier is solidly frozen to the ground and all glacier motion takes place within the ice one to 30 ft above the ground (Goldthwait, 1957). But the Ohio glacier was just like warmer Alpine glaciers, in which as much as 50 percent of the motion has been shown to be due to slipping at the very sole of the glacier (Sharp, 1954). It is logical to think of water lubricating this base, first because evidence presented later shows that the ground was not frozen; so, internal earth's heat may have melted the sole. Second, meltwater is heavier than ice and in these warm marginal areas it could have forced its way hydrostatically from crevasses to the basal ice. Third, great rivers issue even in winter from under the ice cap in south Greenland; so, melting at the base must be continuous.

The source of most of this ice lay northeast of Lake Erie and north of Toronto, perhaps near Lake Simco and the Quebec-Ontario boundary or a little farther north. Between two and 25 percent of the pebbles counted in till of western

Ohio are crystalline quartzites, granites, diorites, schists, most of which can be found in this Ontario area. Among the sedimentary rocks in the till of western Ohio are persistent black limestones (one percent) which match the Ordovician belt north of Toronto. We know of no rocks yet which require ice flow from much farther north. Dreimanis and Terasmae (1958) show that the first motion at Toronto (their "middle till") was westward along the axis of Lake Ontario; so, the ice to the north of Lake Erie shunted southwestward into western Ohio.

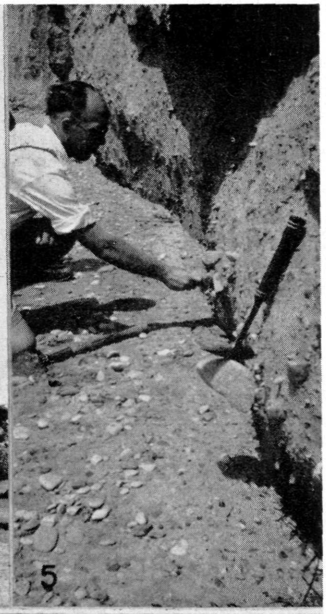


FIGURE 4. Present-day spruce forest near Goose Bay, Labrador, like those in Ohio when the last ice invaded.

FIGURE 5. Glacial till made of boulders, stones, sand, silt, and rock flour and exposed along the north shore of Lake Erie by waves. Pebbles in this till have preferred orientation or "fabric."

FIGURE 6. Glacial till carried at the base of an outlet glacier (South Twin) of the Greenland Ice Cap. A stream has undermined this ice exposing dirt along shear planes over a mound of till already deposited.

These suggest that ice probably started across Lake Erie as a sheet of grounded ice, perhaps 600 ft thick like parts of Ross Ice Shelf in Antarctica.

One of the most obvious features is the lobes into which the ice flow was divided by high bedrock areas such as the Bellefontaine Outlier. This was recognized by the earliest workers (Chamberlain, 1882) from the outlines of terminal moraines (fig. 7). The striae suggest one main stream, the Miami lobe, down through Miami and Montgomery Counties, and another, the Scioto lobe,

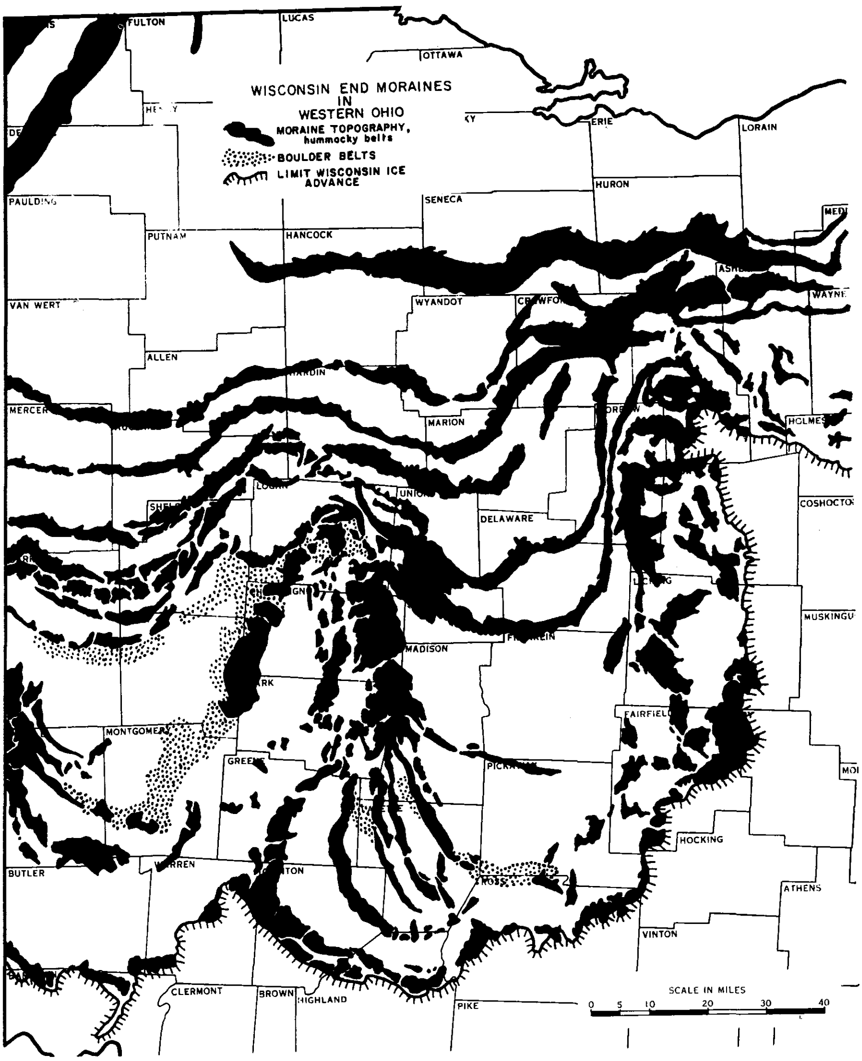


FIGURE 7. Wisconsin end moraines in western Ohio.

down through Delaware and Franklin Counties. The intervening bedrock area above 1,300 ft was covered with thinner ice and probably the streaming was evident on the snow-ice surface as it is in Antarctica today (Apfel, 1952). Later, when the glacier melted away, these marginal lobes became even more deeply lobate.

Now we have good clues as to how fast the ice spread over Ohio. Burns

(1958) estimates from the gradual decrease of ring width over the last 70 yr of the life of buried logs at five localities, coupled with observations in British Columbia showing that trees up to one-half mile away are affected by advancing mountain glaciers, that the ice edge advanced at a rate on the order of 38 ft per year in Butler and Ross Counties. Carbon[14] dates of these same buried logs afford an entirely independent estimate of advance; in lower Scioto lobe, these make the advance average 55 ft per year, and in Miami lobe it averaged 108 ft per year. Advance from the lake near Cleveland to central Ohio was much faster, averaging near 390 ft per year to Sidney and 310 ft per year to Columbus (Goldthwait, 1958).

Glacial studies suggest that these rates are reasonable but do not prove them. Since most glaciers have been in retreat for the last half century, it is hard to find any that have advanced consistantly for many decades. Taku Glacier in a broad valley in southeast Alaska has extended 350 ft per year since last century (fig. 2); until the last decade this glacier ended in water, which may have speeded its advance. North Ice Cap, north of Thule, Greenland, has advanced only two ft per year over patterned ground for some time (fig. 8). Vegetation 100 ft back under the ice was less than 200 yr old (W–532), suggesting only one-half foot gain each year. The Vatnajokull in Iceland is reputed to have expanded up to one mile in a century (Thorarinsson, 1943). At least these rates and dozens of similar ones in modern valley glaciers (Field and others, 1958) show how reasonable the Ohio figures are.

TABLE 2

Various Wisconsin datings suggested for central United States

Radiocarbon years ago	Iowa (Ruhe, Rubin and Scholtes, 1957)	Illinois-Indiana (Horberg, 1955)	Illinois-Indiana (Hough, 1958)	Ontario (Dreimanis and Terasmae, 1958)
Before 38,000			Farmdale Iowan	early Wisconsin
Before 30,000	Iowan		Bloomington	
28,000				
26,000		Farmdale	Cary	
24,000	Farmdale	Iowan		
22,000		Tazewell		main Wisconsin
20,000				
18,000				
16,000	Tazewell	(Brady)		
			Port Huron	
14,000	Cary			
		Cary		
12,000				
	(Two Creeks)			(Two Creeks)
10,000			Valders	Valders

Correlation with adjacent states.—The gradual progress of the ice edge southward across Ohio does not correlate directly at first glance with C^{14} dated advances and retreats to the west (table 2). The total advance occurred between 24,600 and 16,600 yr ago, if outside C^{14} dates in Ohio are reliable; within this period occur all of the classical substages in Illinois (Horberg, 1955) and most of them in Iowa (Ruhe, Rubin, and Scholtes, 1957). It is evident from table 2 that Farmdale advance or Iowan, plus Tazewell and its late phase, Bloomington, were all equivalent to this one advance over Ohio; it is logical that the interstadials represented some halting and even recession in Ohio too. Ice was just moving into Ohio 25,000 yr ago and this tallies with some Iowan dates in Iowa. Fossil molluscs in the windblown loess below the lake beds at Cleveland (Leonard, 1953) suggest a similar correlation for they are like Iowan types in Kansas and the Mississippi

Valley. So far the longest gap in C^{14} dates is 1,600 yr (24,600 to 23,000) and at the end ice was in central Ohio. The slow push to outer Wisconsin terminal moraine in Ohio correlates best by C^{14} with either Bloomington Moraine (late Tazewell) or earliest Cary, depending upon a precision which the C^{14} dates do not afford.

Climax of Glaciation

Different dates on different lobes.—The several lobes in Ohio reached their maximum positions at dates 1,400 yr apart. The Wisconsin Glacier was ploughing logs into the terminal moraine in Miami lobe as far back as about 19,500 yr ago (W–304 Westchester and W–724 or W–738 Hamilton) whereas the ice of Scioto lobe emplaced logs in end moraine 18,100 yr ago (Y–448 Cuba, W–91 Chillicothe, W–331 Anderson). The standard error in these figures ($\pm$400 yr) does not account for this difference. In addition, the moraine pattern (fig. 7) suggests that Scioto lobe moraines in Warren County cross Miami lobe moraine trends; so, the outermost push of Scioto lobe came later. This means that some interlobate areas may have been invaded first from one side, then from the other; in Greene County there are glacial striae both from the northwest and the northeast on the same ledges.

This shifting flow is analogous to recent alternations in Greenland. A few centuries past, the main Greenland Ice Cap advanced several miles overland covered by well-formed patterned soils. Now it is receding, but the contiguous small North Ice Cap is advancing into the same intervening area. Why? The best guess is that sea surfaces, recently freed of summer ice by a rise in temperature, supply more moisture for snow on the coastal North Cap (Goldthwait, 1957) than did pack ice.

Sources of ice supply and motion.—How wide was the zone of bare ice on the Wisconsin Glacier in which all winter snow melted each year? Ahlmann (1939) found that modern Vatnajokull in the cool wet Atlantic atmosphere has an accumulation limit some 16 mi back from the southern edge. In south Greenland, although studies are few, the snow melts for at least 45 mi up onto the ice cap (Holmes, 1955). These should be a minimum measure for the still bigger Wisconsin ice sheet at its southern limit. One might postulate a 50-mi wide zone of completely melting snow so that exposed glacier ice stretched from Chillicothe to over Columbus each summer and from the northern suburbs of Cincinnati to over Dayton, Ohio (fig. 9). Over the rest, snow accumulated faster through the year than it was melted in summer. One main center of gathering snow may have been the ice surface over Lake Erie Basin.

This does not mean that ice did not gather and flow at the same time from eastern Ontario. In fact, the small percentage (ten percent) of crystalline pebbles and minerals at all levels in Ohio till (and a much greater percentage of fines known in Indiana tills) certifies that some ice did come from there. As for Antarctica and Greenland today, there are main streamlines of flow outward from central parts of the ice sheet; on these are superimposed the local fluctuations of snowfall and drifting as surface ice creeps forward at one to 30 in. a day (Schytt, 1955; White, 1956). Motion measurements in Greenland are so slow that if applied to the Ice Age snow falling in Quebec this ice would never have reached Ohio by the demise of the Wisconsin Glacier. Evidently it moved faster in main streams. Snow landing over the Erie Basin took less than 1,000 yr to reach the heavy melting zone near Columbus or Dayton and moved at least two ft a day on the high surface in order to advance the ice margin 350 ft per year even while it melted.

This is not entirely speculation. Careful studies of the deep grooves in Ohio's bedrock and prominent crag and tail (fig. 10), which certainly require many abrading tools and many miles of ice-flow to shape them, indicate that these are concentrated under the former center of each ice lobe and flare outward only slightly to the very outermost terminal Wisconsin position. The grooves on

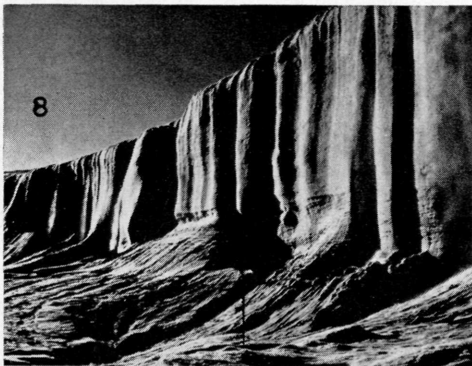
Kelley's Island imply extremely fast flowing ice. In other words, deep abrasion was accomplished where the ice was deepest and flowed farthest. Sharply radiating fine striae were made during thinning ice of later stages for, at the margins of lobes, these aim across the high parts of deep engravings and locally are perpendicular to recessional moraines.

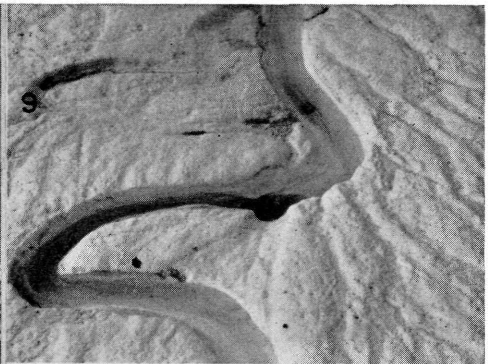
Depth of the ice.—This has always been a fascinating unknown. As our information about Greenland and Antarctica grows, more analogies become available. In Antarctica, over hilly surface as rough as eastern Ohio, the snow surface is about 1,500 to 4,900 ft above the edge 50 mi back from the sea at a dozen places (USNC–IGY Map of Antarctica, 1958). In Greenland snow 50 mi back is 3,000 to 5,000 ft above the ice edge (fig. 11) but the land is more mountainous (Victor, 1953; Holmes, 1955). Vatnajokull, Iceland, lies on low hills comparable to Ohio and it is 4,400 ft high within 50 mi of the ice edge. As much as 200 mi up onto inland ice in Greenland and Antarctica the surface averages 6,400 ft above the ice edge. Until more comparable cases are available, we might estimate 3,000 ft for the depth of ice formerly over Columbus or Dayton, and at least a 5,000-ft depth of ice over Lake Erie Basin.

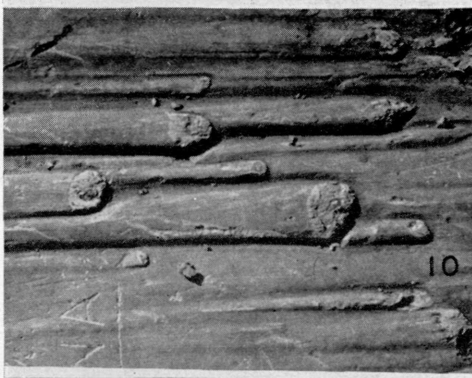
One direct method tells us a little of the steepness of the ice edge itself. All glaciers fit themselves by solid flow to the topography (fig. 12). In the hills of the escarpment in Ross, Fairfield, and Licking Counties, and in Northeastern Ohio, small ice tongues projected between the hills and far down the larger valleys. Detailed mapping of soils along some hillsides suggests that the slope of the ice edge was at least 100 ft per mile in some of these bulging marginal areas (fig. 13). Of course, this is a measure of the extended ice edge against the hills; the ice itself rises more than 200 ft in the first quarter mile from the margin in every active modern glacier.

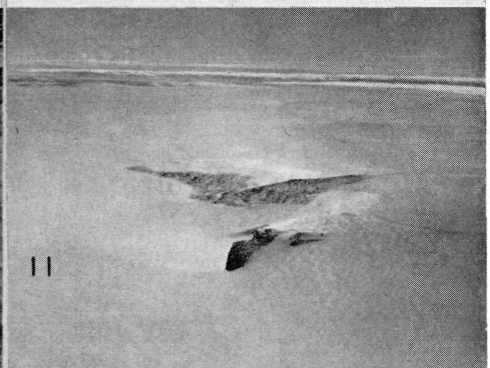
It has long been the hope that the dual compression effects of ice load on top of till and the wetting and drying which compact till could be separated, so that depth of ice load might be calculated. This has not been accomplished yet (Harrison, 1957). Instead, study of the drift sheet does indicate something as to the depth of glacial erosion into the preglacial surface. Since all the loose soil, rubble and rock which were scraped off under the ice, was carried just a little way before redeposition, we can make a direct calculation. Stone counts and sand counts show average travel of two to five mi from the source area and the matrix seems to come from much farther. The average drift depth is 51 ft on the northern uplands (summary of 2,800 wells: VerSteeg, 1933). Detailed recent groundwater studies in western Ohio suggest similar depths for the most recent glacial drift (Water Division, Dept. of Natural Resources, county reports). If we assume, then, that enough extra drift was added through overall southward motion to offset the losses of fines which water carried away, then the ice left a net product about 50 ft deep. One study shows that in Minnesota some 40 percent of the till may be already stained mineral matter (Kruger, 1937). In other words, this was preglacially weathered soil which the glacier scraped up. If Kruger's figure is valid in Ohio, the drift would constitute an original soil layer not over 20 ft deep. If we reduce the remaining 30 ft to original rock density, it was an additional layer of solid rock 25 ft thick.

The change which turned the ice back.—What happened to the plants and animals all this time? Obviously in the area of the ice they were annihilated or driven south. On a botanical basis, Braun (1928) has shown that many which are not easy travelers (do not spread), like *Sullivantia*, must have lingered in protected colonies (refugia) on favorable southerly hillslopes near the ice—otherwise, why should they be clustered near the edge of Wisconsin till now that the ice is long gone? Thomas (1951) has shown that some salamanders and Orthoptera have ice-edge distribution and have moved northward only since glaciation. This is vigorously opposed by Deevey's concept (1949) of a wide tundra belt and ice age animals driven south to the Gulf Coast. However, no bog yet studied in Ohio









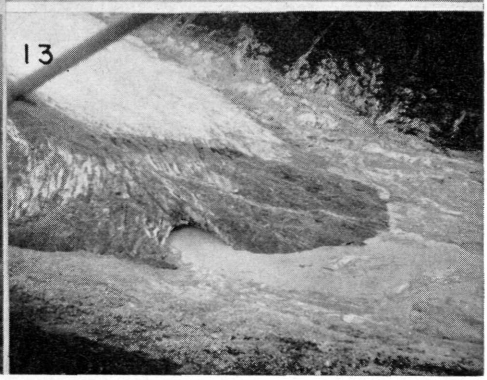


FIGURE 8. Cliff of slowly advancing ice on the east edge of North Ice Cap, Red Rock, Greenland. The ice may have looked like this, advancing over Ohio.

FIGURE 9. Exposed glacier ice where all winter snow melts off near the edge of the Greenland Ice Cap north of Thule. Rivers of meltwater cut the surface as they did in southern Ohio in Wisconsin time.

FIGURE 10. Prominent crag and tail ridges developed beneath the deep ice by abrasion across limestone ledges north of Columbus, Ohio. "Crags" are chert nodules in Columbus limestone. Photo by R. A. Keen.

FIGURE 11. Thirty miles out over the Greenland Ice Cap from Dundas. The ice surface over north central Ohio once looked like this except that no mountain peaks projected through.

FIGURE 12. Typical valley in the edge of the hilly parts of southeastern Ohio. Narrow lobes of ice projected a short way into each valley.

FIGURE 13. Modern glacier terminus, Ferebee Glacier in southeastern Alaska, projected down its valley as a lobate end much as the Wisconsin ice edge did in Ohio.

shows anything but tree pollen dominant at the base; so, I must believe that the spruce and hemlock grew in groups right at the ice edge and moved back over glaciated western Ohio as the ice left. That some mastodons wandered south to Louisiana for the winter is perfectly all right too!

Geological evidence indicates that the last ice sheet did recede in the face of a very mild climate. Subarctic tundra is underlain by permafrost; in patches it even comes into the wooded fringes. Wherever there is permafrost, patterned ground develops (fig. 14), and this in turn involves formation of vertical ice wedges and overturning by frost in the summer thaw zone (two to five ft deep). It is estimated that, in the 15 counties of western Ohio now mapped in detail, some 3,000 road, creek, and pit exposures of soil were inspected. Among these only one showed good possible convolutions (ice-twisted soil layers) and less than three had possible vague wedge structures. This indicates that permafrost was rare if it occurred at all; the nearest well-known obvious large late-glacial patterned structures are fossil forms at 5,000 ft elevation in New Hampshire and

FIGURE 14. Polygonal pattern of sorted soils accomplished by frost action in the thin upper "thaw zone" on top of frozen ground. Nunatarssuaq, Greenland.

New York (Antevs, 1932; Goldthwait, 1940). Lesser effects of intense seasonal frost action is observed on 2,000-ft Pennsylvania hills (Denny, 1956), and in all these cases the glacier ice was nearby. Deducing that permafrost conditions did exist nearly down to 4,000 ft, the mean annual temperature on Ohio's 1,000-ft high land was about 42°F, which is only 11° cooler than today.

By analogy again, what conditions have been required to drive back large modern glaciers during the last half century? In southern Iceland, near sea level, mean annual temperature is 39°F, and summer mean is 51°F. At Juneau, Alaska, nearest station to Glacier Bay, where glaciers have thinned a thousand feet, summers average 56°F. By analogy I believe that a temperature of at least 56°F prevailed during each July in Ohio to drive the ice out. (Winter does not matter for it was too cold in any case!)

Disappearance of The Ice

Decaying of the ice edge.—A glance at glacial maps of Ohio will show that the marginal area, 10 to 20 mi wide, of this last Wisconsin ice sheet and especially the interlobate areas are characterized by ice-contact washed deposits (fig. 15). What does this mean?

Vast areas of similar deposits have been observed in the making today wherever two conditions are met: (1) rapid down melting and thinning of glacier ice, such that its flow is cut off or very sluggish; and (2) presence of hilly topography to cut off the flow and leave great spoon-shaped masses of detached ice. This second condition is met most of the way along the last ice border because it reached into the hills south and east of the Appalachian Escarpment (fig. 15). There are great concentrations of ice-contact deposits south of the high bedrock and interlobate area of Logan County. We can deduce that the last ice sheet began to disappear by rapid melting all over the surface and this produced wholesale thinning of the

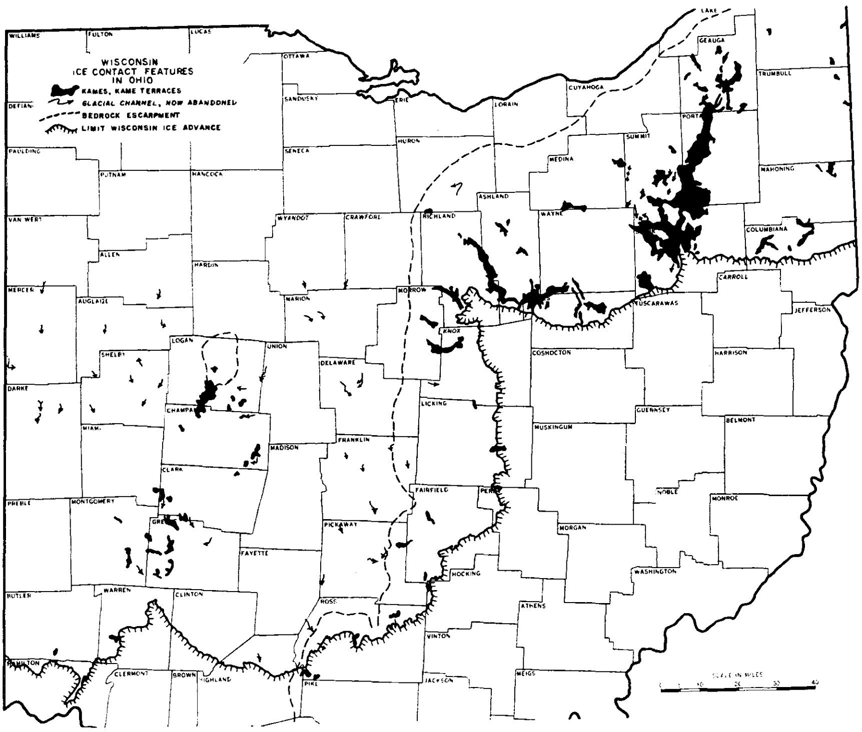
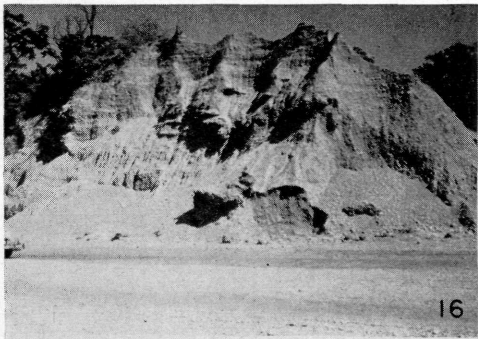


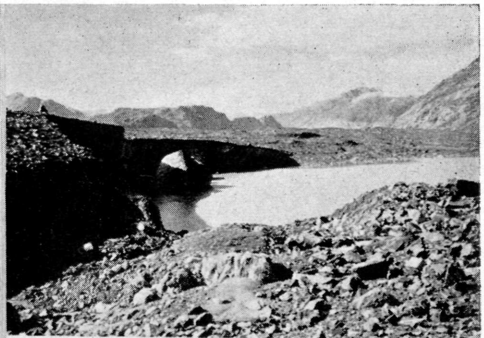
FIGURE 15. Wisconsin ice contact features in Ohio.

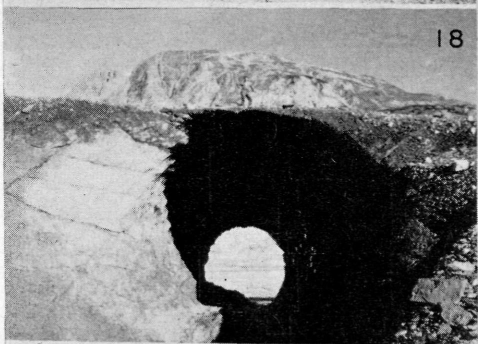
edges. Hills protruded near the edge, and thin ice became covered with dirt which had been near the base.

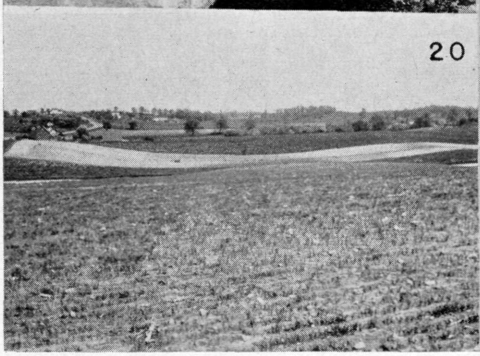
The outstanding deposit of such areas is the glacial *kame*. It is an irregular hummocky mound of layered sand and gravel—some coarse, some fine. Some layers are dirty (i.e., silty), but most layers have the silt and clay washed out. Huge boulders may be sprinkled through the deposit like raisins in rice pudding, and not uncommonly there are irregular blobs of till (fig. 16).

In four large valley areas where this is happening in Glacier Bay today, the dirt does not collect on the ice surface until the ice is nearly gone, for dirt is carried in the lower hundred feet of ice. Finally it collects from two to four ft thick on the last ice surface and protects it from melting for many years (fig. 17). In the meantime, meltwater and rain open a honeycomb of passages through the hidden blue ice beneath. Collapse of these passages and enlargement of the surface entryways start the surface dirt sliding and washing into the pits (fig. 18). This develops large circular sink holes with pools in the ice. Floods of meltwater collect washed and half-washed sand or gravel at the bottom (fig. 19). Eventually the surrounding ice melts away and the deposit at the bottom of each former pool is left as a mound. Where decaying ice was previously high, there is now a trough; the topography of the decaying ice is preserved in reverse (fig. 20). Where









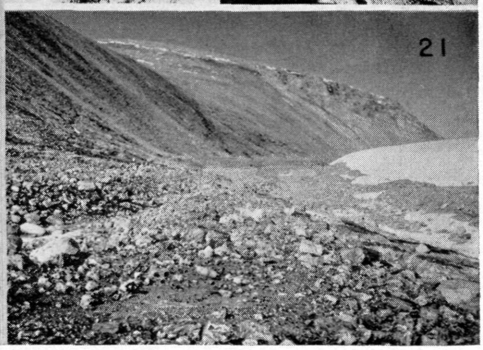


FIGURE 16. Pit excavation in glacial kame deposits of Southern Hills, Dayton, Ohio. Sandy gravel with cemented masses and boulders.

FIGURE 17. Remnant ice mass from decaying glacier covered with soggy dirt and accompanied by meltwater lakes which undermine the ice cliffs. East side, Muir Inlet, Alaska.

FIGURE 18. Decaying ice mass with section of old stream tunnel and cover of dirt. Muir Inlet, Alaska.

FIGURE 19. Funnel-shaped pits developed in dirt-covered glacier ice near the end of Tasman Glacier, New Zealand. In these pools sorted sandy gravel gathers to become kames.

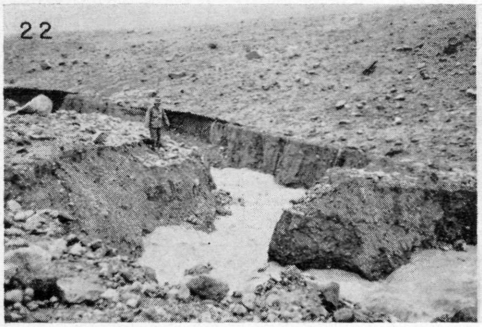
FIGURE 20. Kame mounds of sandy gravel south of Akron, Ohio. These were deposited in summer in pools surrounded by dirty glacier ice.

FIGURE 21. Kame terrace built recently by streams depositing gravel between the ice on the right and the hillside on the left. Nunatarssuaq, Greenland.

the last buried ice slabs slowly melt out, funnel-shaped *kettle holes* develop in the gravels (fig. 30). Kames and kettles come together.

Sometimes a flat-topped shelf or terrace of sandy gravel lies on a hillside above a valley partly filled with kames. These are kame terraces formed in the early stages of kame accumulation where tumbling meltwater washed its full load into the trough between the down-wasting ice and a hillside already there. Later, when the ice melted, a shelf was left, complete with kettle holes and sloughed boulders, high and dry along the hillside (fig. 21). These are found in main valleys through the hills, especially in northeastern Ohio (fig. 15).

On nearby hillslopes associated with kames are short glacial *channels* cut along the hillside rather than directly downhill as any "sensible" stream would go. This is a common feature around decaying ice in hilly Alaska (fig. 22). There may be a dozen or more channels, 100 to 1,000 ft long and five to 50 ft deep, cut in till. It is clear from recently studied examples that dying ice fills the deep valley so that water along the valley wall trims the till away, first at one level and then at a lower level. No characteristic series of channels has yet been found, even in the hilly northeastern parts of Ohio; but there are many short abandoned channels over the lower plains, and especially through hummocky end moraines where the initial drainage was forced away from present courses by ice openings (fig. 23).



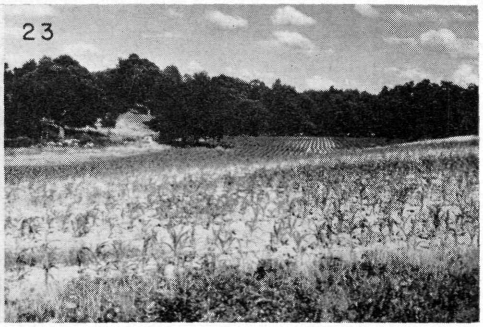


FIGURE 22. Glacial stream flowing along the margin of dirt-covered wasting ice (background). The meltwater is cutting a channel along what will become sloping hillside as the ice melts. Plateau Glacier, Muir Inlet, Alaska.

FIGURE 23. Old glacial channel in Logan County, Ohio. Water once flowed from right to left from the glacier to cut this valley.

Systematic northward melting.—Every large ice cap which has been retreating during the last half century has an extra broad zone of melting ice surface during each summer. In cold polar glaciers, like those in north Greenland, water cannot penetrate the cold ice (14°F) without refreezing; so, it passes over the surface in rivers. In lower latitudes, deep ice is just about at melting temperature. Here floods of meltwater pour into holes one to ten mi back from the margin. Thus, in retreating, the ice thins, melts from within, and melts back at the edge faster than motion supplies new ice to the edge. (Recent tunnel measurements show that small open tunnels under 100 to 200 ft of ice cover close at one-half to one ft per year from all sides and bigger rooms close at two or three ft per year [Rausch, 1958]. Water must reenlarge these natural tunnels each year.)

The only record of this sort of ice tunnel in Ohio is the random *esker* ridge which ends in or near a belt of end moraine (fig. 24). In living glaciers, passages in the upper ice are almost free of dirt, for all the dirt load of a glacier is concentrated in the lower ten to 200 ft of ice. Where gradient is low, it favors deposition of alternating layers of sorted sands and gravels. These deposits may get deeper until the tunnel floor is built up with a ribbon of gravel ten to 100 ft thick. Not all the streams in this dirty ice at the bottom of the glacier are depositing, either, for some cut channels. For this reason the deposits may not be continuous. (This does not explain all eskers. Eskers in New England which climb up over a divide, or eskers that are continuous horseback ridges 50 mi long must have been made in hydrostatic pipe systems in the ice.)

The eight Ohio eskers are actually chains of ridge segments (fig. 27). We presume that the drainage system was all one long ice tube, since ridge segments lie end on end. Each esker contains layers of good water-washed sand and gravel dipping gently in every direction. Some have several tributaries, as in Delaware County where the esker involves parallel ridges. Another through Pickaway County has patches of till on the crest; these might have dropped from the ice roof. More difficult to explain is the four- to six-ft veneer of ice-laid till on the buried esker east of Columbus. This might mean that it was made under the edge of an earlier ice sheet, or the last ice sheet itself might have slid over its own esker after waters formed it.

A closely related form is the isolated high mound of sandy gravel erroneously referred to as an Indian Mound. However, these are up to a half mi in diameter and 50 to 200 ft high with carefully cross-bedded gravel—busy Indians! Actually, they are ***moulin kames*** standing above Ohio's plains (fig. 25 and 27).

Similar mounds were seen forming on Barnes Ice Cap, Baffin Island. Surface drainage plunges into a huge cylindrical hole half a mile back from the ice edge and over 200 ft deep. Apparently the water column has forced and melted its way out below, but torrents of meltwater swirling into the hole gathered dirt from the basal ice and piled it in the middle of the bottom like sugar in a coffee cup. Ten mi away a similar knob of gravel was melting out so its gravel slumped to an isolated chocolate drop-shaped pile (fig. 26).

Third and most extensive is the ***outwash plain*** or valley train. Wherever a glacier ends in a region which slopes away from the ice, the waters which issue from one or more tunnels in the ice edge head down the nearest valley lowland (fig. 28). Many million gallons are produced on warmest days. Coming from the basal ice, the water is milky with the rock flour of glacial till and it has a full bed load of sand and gravel (fig. 29). In fact, more tons of particles are fed into the stream at the glacier than ever reach the lake or sea; so, there must be constant addition to the surface or the plain. Some load is dropped on each bar, thus filling up the shallow channel and causing the water to spill over to the right or left and start an additional channel. All glacial outwash rivers become braided by dividing into many channels each ten to 100 ft wide and one to ten ft deep. The water runs nearly flush with the surface. Most of the coarsest cobble part of the load gets left on the first few upstream bars, but in their place the water gathers some smaller pebbles already there. There is constant exchange such that where the meltwater reaches its delta, it has a smaller load composed mostly of sand and silt. For this same reason, the gravelly upper part of the plain is steeper (over ten ft per mile) than the sandy lower part. When one side of the valley plain builds ten to 20 ft higher, the braided currents shift to the low side of the valley. The resulting "valley train" is smooth with only minor little scarps (five to 20 ft) and abandoned channels.

This was the scene in most south-flowing valleys in Ohio as the ice retreated (fig. 28). On the east and west walls of each valley, such as the Licking, the Hocking, the Scioto, or Miami River, there are steplike shelves of sand and gravel which at first glance look like kame terraces (fig. 30). Unlike kames, the shelf-tops match across valleys, indicating that they are the remnants of a once-continuous sandy gravel plain. Following glaciation, the meandering rivers have dug down and trimmed away central parts of these smooth glacial valley trains; much of the gravel is gone; yet, it is still our third richest underground resource material.

The glacial map of Ohio shows a large area of outwash from Logan County to Greene County in west central Ohio (fig. 27). This is not only Ohio's biggest outwash, and a superb drought-defeating source of groundwater, but it has no

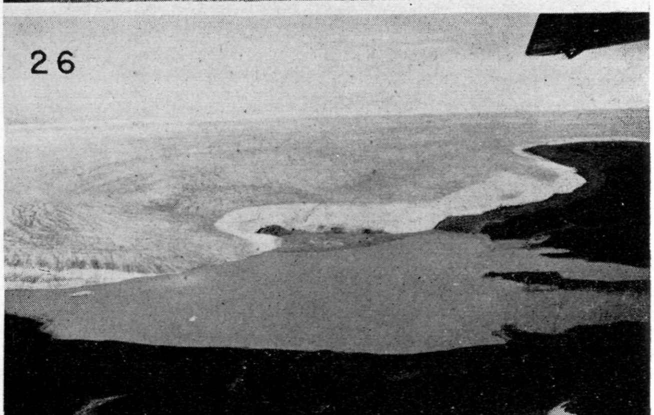


FIGURE 24. Pit excavation follows a winding esker ridge which continues to the right in the distance near New Richland, Ohio. Photo by J. L. Forsyth.

FIGURE 25. Mound of layered sandy gravel, called a moulin kame, formed in a large hole in the former glacier ice east of Urbana, Ohio.

FIGURE 26. Moulin kame mound in the center melting out from the retreating southwest ice edge of Barnes Ice Cap, Baffin Island.

western valley wall! In fact, to explain its course, cutting from Mad River watershed through low hills to Little Miami watershed, we must presuppose an ice wall on the west which prevented its flow into Mad River and Miami Valleys. Verifying this hypothesis, there is ice-laid till pushed up from the northwest onto the high west side of this plain.

A second general feature of these outwashes is that most valleys contain not one but two or more persistent low levels of valley train. (Of course, the highest levels beyond the Wisconsin terminal limit have deep reddish soils and relate to

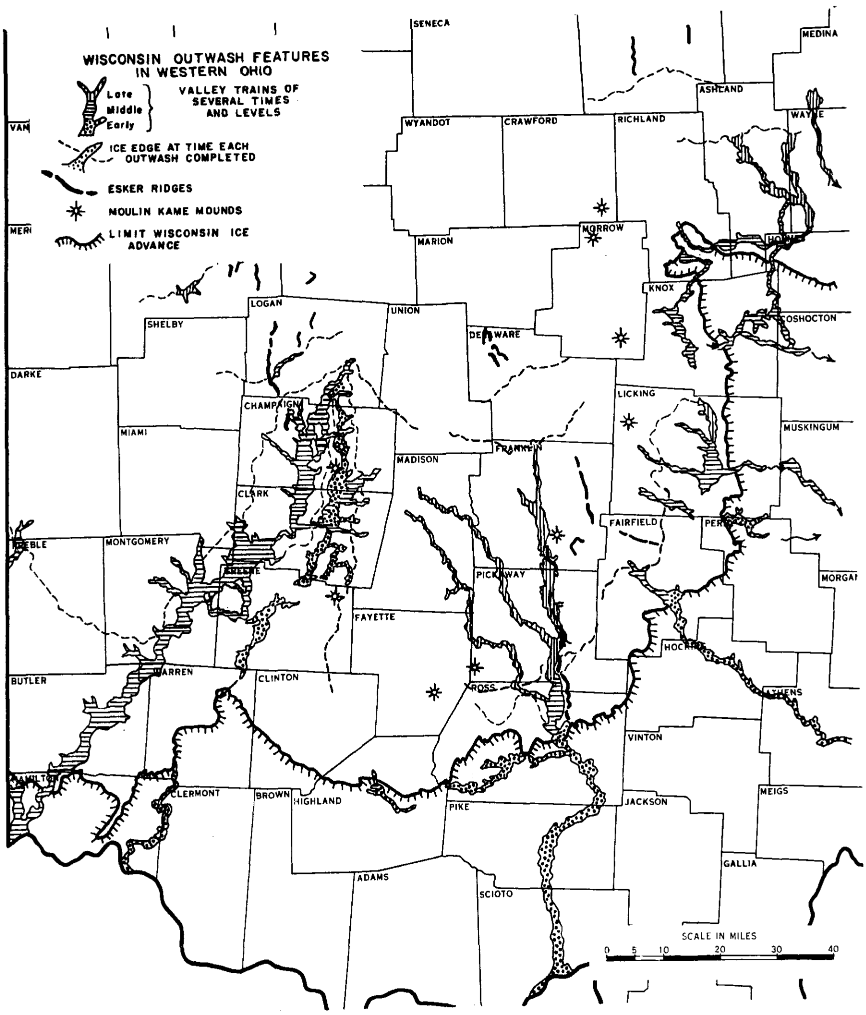
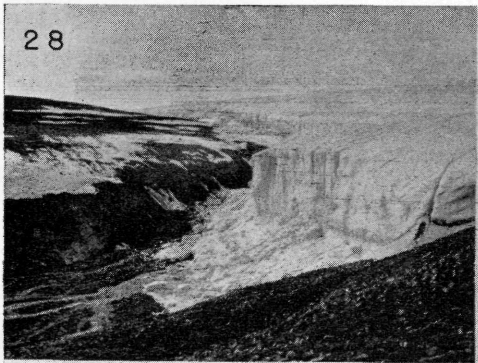
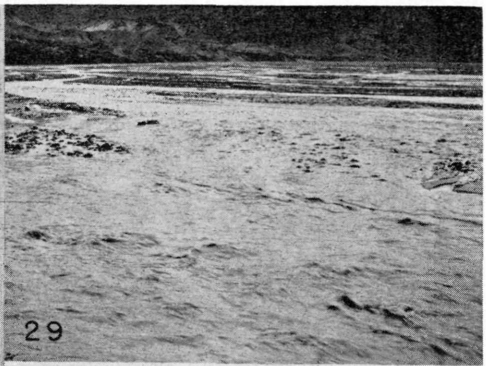


FIGURE 27. Wisconsin outwash features in western Ohio.

entirely different, earlier glaciations.) There are three different lower levels with different kinds of pebbles and different soil cover in Licking, Hocking, and Scioto Valleys (Kempton and Goldthwait, 1959). Longitudinal profiles up and down the valleys depict each level beginning at a different place; yet, each one emanates from a group of ice-contact kames and underdrained kettle holes in end moraine (fig. 31). Clearly the ice edge was wasting away at a different position when each valley train was completed (fig. 27). The higher the terrace level above to-





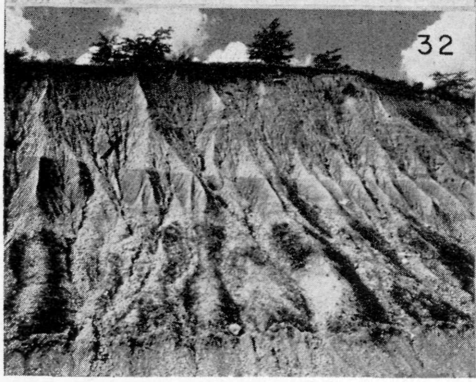


FIGURE 28. Head of outwash plain at the edge of North Ice Cap, Nunatarssuaq, Greenland. Floods of meltwater build up the gravel plain (lower left) as they did once in each Ohio valley.

FIGURE 29. Floods of silty gray meltwater from Adams Glacier in southeast Alaska, spreading over its outwash plain on a warm summer day.

FIGURE 30. Terrace cut out of glacial valley train along the Mad River Valley just east of Dayton, Ohio. This is composed of sorted sand and gravel beds.

FIGURE 31. Kettle hole collapsing where a modern outwash plain is being built from Hugh Miller Glacier, Alaska. As buried ice melts, out pits form in the sandy gravel.

FIGURE 32. Several glacial tills, some rich in clay and some stoney, are piled one on the other and exposed in this railroad cut south of West Liberty, Ohio.

FIGURE 33. Rolling high ground called end moraine. These form belts across the state. This one is in Greene County, Ohio.

day's river, the farther south and earlier the ice edge was when it formed. From the glacial map, one can deduce that the bulk of outwash was deployed down the valleys in the early stages of retreat. When the ice edge lay north of Powell and Union City Moraine line, no significant outwash deposits were made even although some land naturally drains south.

Periods of halt.—The 23 logs actually dated as annihilated during the last ice advance are generally found toward the bottom of a layer of till, but near the surface. In a few cases like that at Sidney (Forsyth and La Rocque, 1956), there are one or two distinctly different layers of till above the logs; some are more pebbly; some vary in color; all vary in stone count lithology (fig. 32). This suggests a change in direction of ice motion after the early advance, or perhaps a retreat and readvance from a different direction.

Halts are most clearly expressed by the belts of *end moraine* hung like festoons across the glacial map in Ohio (fig. 7). Each one is a semicontinuous belt of ground, ten to 300 ft higher than surrounding glacial deposits, or perhaps just thicker till where it is plastered onto bedrock hills as in Logan County or northeastern Ohio. The surface is rolling or hummocky in contrast to the relative flatness of nearby plains, and it bears greater density of scattered boulders (fig. 33). It is composed of till and many lenses of gravel, but there is usually no favored orientation of pebbles or fabric. Thus, each belt of hummocks appears to be dumped or unloaded from the glacier front while it stood along a particular line. Some of these are continuous enough so they can be traced all across western Ohio; others are broken by sharp hills, wide river valleys, or they just fade out. Here and there it is hard to tell which are time equivalents. Most of the moraines have double parallel high areas with minor creeks traveling the line of the moraine between them. Each major moraine appears to be a composite of deposits along closely spaced successive parallel positions of the ice. Where six completely separate moraines converge northward in Clark County, they become the eight-mi wide Cable Moraine belt, yet generally an irregular ridge within the belt marks the position of a contemporaneous ice edge to match each moraine farther south (fig. 7).

No glacier is known today which is producing an end moraine of the breadth and volume of those in Ohio (two mi wide, average 50 ft high). Only in height have they been equalled by the looped moraines built by some mountain glaciers since 1700. On Barnes Ice Cap, Baffin Island, moraines about one-tenth as wide were observed (Goldthwait, 1951). What do they imply in this modern case? Dirt is brought up on shear planes to the ice surface never more than 200 ft vertically above the ice edge nor a quarter mi back from the edge (fig. 34). Here it melts out from June to August and slides down the trickling wet ice surface. It stacks up three to four ft deep over dead ice at the very edge of the glacier, making a temporary ice-cored moraine (fig. 35). Here and there this buried ice melts from below where water opens a hole, where streams cut, or where an ice tube collapses (fig. 36). The covering dirt slips and slides, first one way and then another, until it is all superimposed in heaps on earlier glacial deposits beneath. It ends up dumped as a series of hummocks along the ice edge. If the ice edge is reactivated, it pushes into these hummocks and may repeat over and over, adding to the thickness and width of the accumulated mass (fig. 37). From the sparcely disseminated dirt in this glacier ice (0.3 to 3.0 percent) and sample motion measurements, it has been estimated that the smallest of these moraines takes a century to accumulate.

Each Ohio moraine has easily ten times this bulk, and yet C^{14} dates show that we have nothing like this 1,000 yr to attribute to each belt of moraine. Thus, either ice over Ohio was dirtier than today's example, or the dirt zone was deeper than 200 ft, or the surface ice motion was greater than half a foot per day. Or possibly some moraines were pushed up rapidly. If all three were double, then each Ohio moraine might have been accumulated in a century or two.

Duration of retreat.—The youngest C^{14} samples buried in the outermost Cuba-Hartwell Moraine are 16,560 ± 230 (Y–450) yr and 17,980 ± 400 (W–331) yr old. Next youngest are lacustrine and swamp deposits on top of the Wabash Moraine way up northwestern Ohio, these postglacial burials are 14,300 ± 450 yr old (W–198). If we believe that the ice left the terminal area generally 17,000 yr ago, then we have only about 2,700 yr to remove the ice from over most of Ohio, including the formation of eight moraines to the Wabash Moraine line. Actually, if we assume half of the time of deglaciation was absorbed by halts when the ice edge built moraines, the retreat moved at averages near 300 ft per year, and at least 150 yr was spent by the ice edge at each moraine.



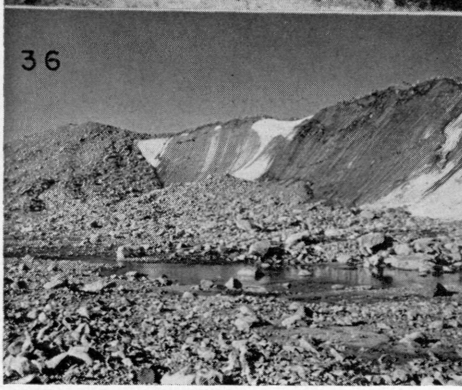



FIGURE 34. Melting ice slope at the south margin of Barnes Ice Cap, Baffin Island. Dirt is brought up only to the white ice lip by shearing upward from the bottom.
FIGURE 35. Dirt (till) accumulating on the ice at the very edge of the ice cap near Dundas, Greenland. This is ice-cored moraine.
FIGURE 36. Till slipping and sliding down the irregular undermined edge of Barnes Ice Cap, Baffin Island, to form hummocks of end moraine in a belt.
FIGURE 37. End moraine ridge with hummocks deposited several centuries ago by Barnes Ice Cap, Baffin Island.

The history of modern glaciers, as in Glacier Bay, Alaska, where new snow supply has been largely cut off since the end of the last century, indicates a maximum recession of about 1,000 ft per year (Field, 1958). This may be the most realistic rate for recession over Ohio, even though there was no water body to accelerate recession of the ice here. Certainly this rate might apply to the ice disappearing up the Erie Basin when it was filled with glacial lakes.

Readvances or long halts?—One reason for insisting that the end moraine deposits represent most of the time of overall retreat is that there is evidence along at least three moraine fronts that a significantly long time was involved, or there was actual reexpansion of the ice. Within historic times in the Alps, Norway, and Iceland (Ahlmann, 1953), glaciers have been notorious for pulsating readvances; so, while continental glaciers must work sluggishly and respond only to long-continued climatic changes, such fluctuations are still to be expected.

The first evidence of long halt is the approximate northward limit of a substantial cover of windblown *loess* on top of Wisconsin drift. This is near the outer edge of Camden Moraine and the Reesville Moraine (dashed line in fig. 38). In several dozen samples of soil, two to five ft deep, selected for its smooth silty character, 20 percent was clay (less than 20 μ) derived largely by soil forming

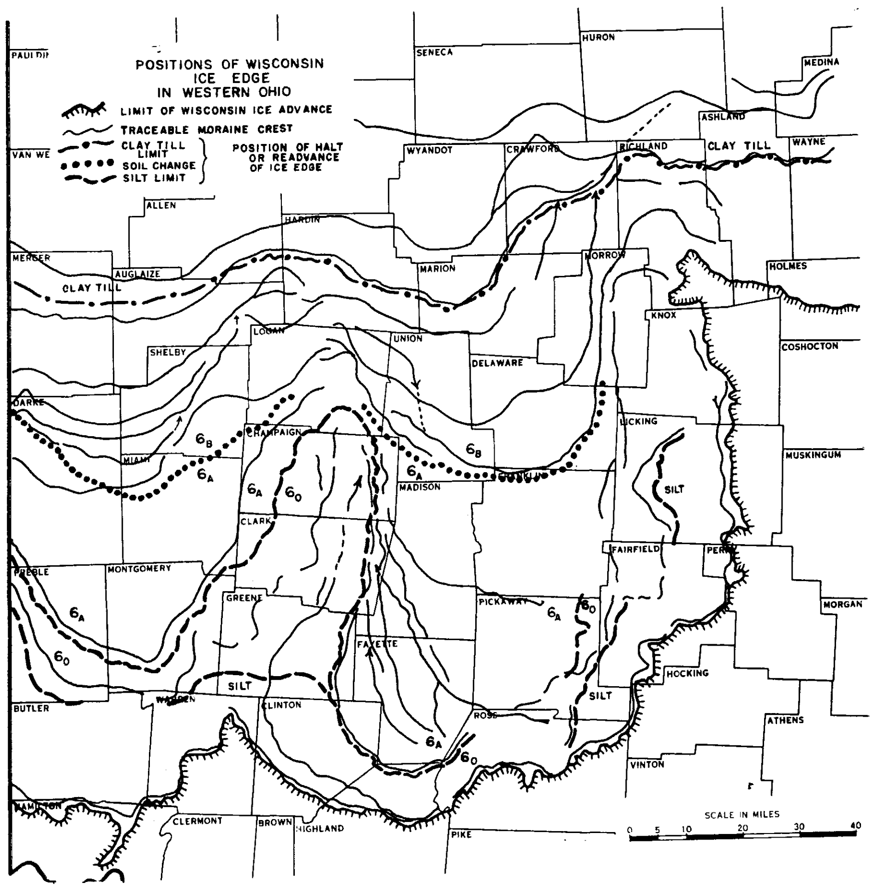
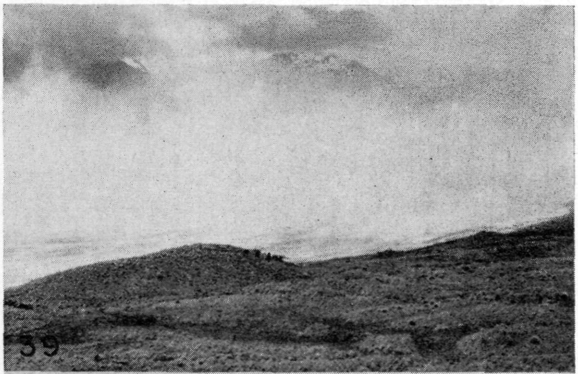


FIGURE 38. Positions of Wisconsin ice edge in western Ohio.

processes, 68 percent was silt (up to 0.05 mm), and only 11 percent was sand, mostly fine. Through wide areas to the south, this covers the gritty-pebbly till, but only thin irregular patches occur north of this "silt" line. Such loess deposition is observed and measured wherever great outwash "valley trains" are built today, as in Greenland or New Zealand. Summer floods spreading over the plains between multitudes of channels leave patches of silt. The silt is picked up and dusted over the leeward countryside by the next gusty frontal wind (fig. 39).

The predominance of minerals, quartz, felspar, and calcite, in Ohio silts is like that in Ohio outwash; so, there is probably an association.

This silt limit is near, but not identical with a change in soils type temporarily designated "Miami 6O" vs. "Miami 6A." Actually, the "6O–6A" line is based on other characteristics such as the greater depth, yellow-red color, and the marked structure in the 6O soils. However, the two limits (silt and soil) are fairly close



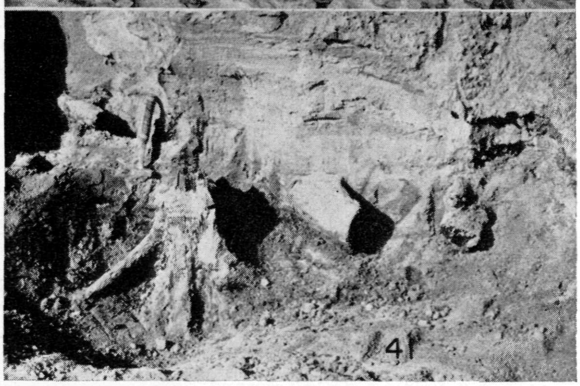


FIGURE 39. Clouds of silt-size dust (loess) being blown up by fresh frontal winds off of an outwash plain in Tasman Valley, New Zealand. The silts were deposited by recent floods of glacial meltwater.

FIGURE 40. Ice cliff of Barnes Ice Cap, Baffin Island, advancing into a small ice-dammed lake much as happened many places in northern Ohio.

FIGURE 41. Buried forest stump which grew on the clay beds of former glacial lakes at Castalia, Ohio. This one is 8,500 yr old, and was buried by marl.

except in the interlobate area from Greene County north to Logan County. One might argue that there was no dust available in this interlobate area; yet, there was more glacial outwash in Mad River Valley than in the silty areas of Warren, Clinton, and Highland Counties. The Mad River valley train (fig. 27) leads back nearly to the Farmersville Moraine on the west, and to the Reesville line traced into Cable Moraine on the east. This shows that these two were indeed simultaneous ice edges. Lack of silt cover here is a dilemma but the long halt signified by some two ft of loess to the south suggests one or two long halts.

A second soils change occurs in the vicinity of Powell and Union City Moraines (dotted line in fig. 38). Schafer is finding by a study in progress a subtle change from silty till with sticky brown clay subsoil, and topsoil over ten in. deep (Miami catena) northward to a clay till with shallow topsoil, poorer structure, darker B zone, and lower calcium carbonate (Morley catena). Wenner (thesis in press) has shown that clay content in the parent till increases from about 21 percent to about 33 percent in passing north over this line. Since climate, weathering, and biological activity are identical, the change must relate to the difference in original composition, but this difference reflects new source materials, new directions of ice motion, and therefore probably some readvance. It coincides in position with one of the most prominent end moraine systems in Ohio, and with the last of the valley train systems in Scioto lobe.

The third significant line is either the Wabash Moraine or some line just north of it (dash-dot line in fig. 38). The Wabash Moraine itself cuts directly across earlier moraines, overriding the Powell, Broadway, and St. Johns Moraines in Crawford County. It is much more bouldery than areas to the south, and Gregory (1956) found that soluble carbonates increase from a few percent to over 17 percent in this moraine. Another increase in clay content in till was detected at the south edge of Huron County (Campbell, 1955). Pedologists like Holowaychuk recognize gradual increase of clay northward in the Miami lobe, most markedly between the Wabash and Defiance Moraines. This is attributable to lake floor clays and silts picked up by a readvancing ice (fig. 40). Now we are beginning to find these buried lake floor sediments themselves beneath the uppermost till sheet in Huron County (Campbell, 1955), and Seneca County (Coash, personal communication). Very clearly the ice retreated far into the Erie Basin in what Dreimanis (1958) calls the Erie Interstadial. A lake was trapped in the area between the ice and the Defiance Moraine, and then the ice readvanced nearly to the Wabash Moraine. There is some likelihood that Defiance Moraine is principally an earlier (Iowan?) moraine overriden at this time. Since the oldest logs found buried in late glacial sediments on top of the Wabash Moraine are 14,300 ± 450 yr (W–198) old, it looks as though this readvance culminated at least 14,500 yr ago.

To the west this is the time equivalent of the Cary substage according to one recent estimate (table 2, Ruhe, Rubin, and Scholtes, 1957). In northeastern Ohio this connects to the "late Cary" readvance demonstrated so neatly by a clay till (30 to 50 percent clay) lying over a silty-clay till (20 to 30 percent clay, Shepps, 1953). North of Lake Erie, Dreimanis (1958) has shown the same clay-rich till layer over a silty till, which was actually brought by ice moving northwestward out of the Erie Basin. With such localized radial movement, he and Holmes (1952) suggest a late center of motion right over the Ontario Basin. This is good reason to think that even the Erie Basin was still a local ice-forming center and together with Ontario Lake Basin, it developed the final push of ice over Ohio 14,500 yr ago.

Aftermaths of Glaciation

Aftermaths of glaciation.—By 13,000 yr ago (W–33, Y–240, S–25), several glacial lake levels had existed in the Erie Basin. Niagara outlet was well blocked by ice, and water standing 230 ft deeper than present Lake Erie first spilled into the Wabash River at Fort Wayne (the Maumee stage). Oscillations produced several levels well worked out by Carney (1909, 1910, 1913). During the short, slowly-dropping level called Lake Akona, the water body extended north to the thumb of Michigan as ice retreated, and it drained by some lower outlet. This long warm period gave trees a chance to come in right down to the lakeshore for, when the lake rose to Whittlesey stage by resurgence of ice (Port Huron Moraine, Michigan), the rising waters drowned the forests and buried the trees under Lake Whittlesey shore sands. This is the earliest firm glacial lake dating.

It is notable that even while ice was as near as northern Lake Erie (Port Huron Moraine), spruce trees grew next to early Lake Whittlesey. Bogs studied by Sears (1930) in northern Ohio show that the accumulation started first with dominance of spruce and some fir pollen. Possibly during the retreat from the Wabash Moraine to northern Lake Arkona between 14,300 and 13,000 yr ago, there was a period of treeless plains, for Lake Maumee has produced no logs. This was a narrow barren belt, if it existed, for trees grew on Wabash Moraine at the start. Thus, rapidly did spread the groups of spruce trees and with them moved the forest-loving mastodon and giant beaver. These populated the country widely for they are associated with the forest in bogs at Columbus and West Jefferson (Thomas et al., 1952). One is dated near 11,480 yr ago (Y-526), which is Two Creeks interstadial time farther west, and the other is 9,600 yr ago which is the end of Valders time.

In another thousand years, however (8,500 yr ago), warm climate had come to Ohio, for Castalia Bog (fig. 41) at this time preserved fragments of oak, elm, poplar, ash, maple (Burns, 1958). Deer became plentiful (Campbell, 1955), but the mastodon was fighting a losing battle. Mean annual temperatures in the 50's (°F) were upon us. Water level dropped to near the present Lake Erie level and the ice had faded rapidly into eastern Canada (Hough, 1958). The ice age was gone from Ohio.

REFERENCES CITED

Ahlmann, H. W. and **S. Thorarinsson.** 1939. Vatnajokull, scientific results of the Swedish-Icelandic investigations 1936–37–38, Chapter 6. The Accumulation. Geografiska Analer 1: 41–66.

———. 1940. Chapter 10. The relative influence of precipitation and temperature on glacier regime. Geografiska Analer 3–4: 188–205.

Ahlmann, H. W. 1953. Glacier variations and climatic fluctuations. American Geogr. Soc., Bowman Memorial Lectures, Series 3.

American Geographical Society. 1958. Antarctica. (map) 1: 6,000,000 prepared for National Acad. Sci., U. S. National Committee, I.G.Y.

Antevs, E. 1932. Alpine zone of Mt. Washington Range. Auburn, Me.: 118 pp.

Apfel, E. T. 1952. Antarctica drainage pattern (abst.). Bull. Geol. Soc. Amer. 63: 1232.

Braun, E. L. 1928. Glacial and post-glacial migrations indicated by relic colonies of southern Ohio. Ecology 9: 284–302.

Burns, G. W. 1958. Wisconsin age forests in western Ohio, II. Vegetation and burial conditions. Ohio Jour. Sci. 58: 220–230.

Campbell, L. J. 1955. The late glacial and lacustrine deposits of Erie and Huron counties, Ohio. The Ohio State University, Ph.D. dissertation.

Carney, F. 1909. The raised beaches of the Berea, Cleveland, and Euclid sheets, Ohio. Denison Univ. Science Lab. Bull. 14: 262–287.

———. 1910. The abandoned shore lines of the Oberlin quadrangle, Ohio. Denison Univ. Science Lab. Bull. 16: 101–117.

———. 1913. Some proglacial lake shorelines of the Bellevue quadrangle, Ohio. Denison Univ. Science Lab. Bull. 16: 231–246.

Chamberlin, T. C. 1882. Terminal moraine of the second glacial epoch. U. S. Geol. Sur. Ann. Rpt. 1882: 291–402.

Deevey, E. S., Jr. 1949. Biogeography of the Pleistocene, Part 1: Europe and North America. Bull. Geol. Soc. Amer. 60: 1315–1416.

Denny, C. S. 1956. Surficial geology and geomorphology of Potter County, Pennsylvania. Geol. Surv. Prof. Paper 288.

Diamond, M. 1958. Air temperature and precipitation on the Greenland Ice Cap. U. S. Army Engineers Snow, Ice, and Permafrost Research Establishment, Research Rpt. 43.
Dreimanis, A. 1958. Wisconsin stratigraphy at Port Talbot on the north shore of Lake Erie, Ontario. Ohio Jour. Sci. 58: 65–84.
——— and **J. Terasmae.** 1958. Stratigraphy of Wisconsin glacial deposits of Toronto area, Ontario. Proc. Geol. Assoc. Can. 10: 119–135.
Field, W. O., and others, 1958. Geographic study of mountain glaciation in the northern hemisphere. American Geog. Soc. Rpt. to U. S. Army Quartermaster.
Forsyth, Jane L. and **A. La Rocque.** 1956. Age of the buried soil at Sidney, Ohio. (abst.) Bull. Geol. Soc. Amer. 67: 1696.
Goldthwait, R. P. 1940. Geology of the Presidential Range. New Hampshire Acad. Sci. Bull. 1: 43 pp.
———. 1951. Development of end moraines in east-central Baffin Island. Jour. Geol. 59: 567–577.
———. 1957. Study of ice cliff in Nunatarssuaq, Greenland. The Ohio State University Research Foundation, Project 636, Rpt. 17.
———. 1958. Wisconsin age forests in western Ohio, I. Age and glacial events. Ohio Jour. Sci. 58: 209–219.
Gregory, J. F. 1956. The Pleistocene geology of Crawford County, Ohio. The Ohio State University, Thesis for M.Sc.
Harrison, P. W. 1957. A clay-till fabric: its character and origin. Jour. Geol. 65: 275–307.
Holmes, C. D. 1941. Till fabric. Bull. Geol. Soc. America. 52: 1299–1354.
———. 1952. Drift dispersion in west-central New York. Bull. Geol. Soc. Amer. 63: 993–1010.
Holmes, G. W. 1955. Project Mint Julep, investigation of smooth ice areas of the Greenland Ice Cap, 1953, Part II. Air Univ., Arctic, Desert, Tropic Information Center. 100 pp.
Horberg, L. 1955. Radiocarbon dates and Pleistocene chronological problems in the Mississippi Valley region. Jour. Geol. 63: 278–286.
Kempton, J. P. and **R. P. Goldthwait.** 1959. Glacial outwash terraces of the Hocking and Scioto River valleys, Ohio. Ohio Jour. Sci. 59: 133–151.
Hough, J. L. 1958. Geology of the Great Lakes. Univ. Illinois Press: 90–129.
Kruger, F. C. 1937. A sedimentary and petrographic study of certain glacial drifts of Minnesota. Amer. Jour. Sci. 34: 345–363.
La Rocque, A. and **Jane L. Forsyth.** 1957. Pleistocene faunules of the Sidney cut, Shelby County, Ohio. Ohio Jour. Sci. 57: 81–89.
Leonard, A. B. 1953. Molluscan faunules in Wisconsinian loess at Cleveland, Ohio. Amer. Jour. Sci. 251: 369–376.
Leverett, F. 1902. Glacial formations and drainage features of the Erie and Ohio basins. U. S. Geol. Survey, Monograph 41.
Martin, P. S. 1959. How many logs make a forest? Ohio Jour. Sci. 59: 221–222.
Rausch, D. O. 1958. Ice Tunnel, Tuto area, Greenland. U. S. Army Engineers, Snow, Ice, and Permafrost Research Establishment, Tech. Rpt. 44.
Ruhe, R. V., M. Rubin, and **W. H. Scholtes.** 1957. Pleistocene radiocarbon chronology in Iowa. Amer. Jour. Sci. 255: 671–689.
Schytt, V. 1955. Glaciological investigations in the Thule Ramp area. U. S. Army Engineers, Snow, Ice, and Permafrost Research Establishment, Rpt. 28.
Sears, P. B. 1930. A record of post-glacial climate in northern Ohio. Ohio Jour. Sci. 30: 205–217.
———. 1941. Post-glacial vegetation in the Erie-Ohio area. Ohio Jour. Sci. 41: 225–234.
Sharp, R. P. 1951. Accumulation and ablation on the Seward-Malaspina Glacier system, Canada-Alaska. Bull. Geol. Soc. Amer. 62: 725–744.
———. 1954. Glacier flow: a review. Bull. Geol. Soc. Amer. 65: 821–838.
Shepps, V. C. 1953. Correlation of the tills of northeastern Ohio by size analysis. Jour. Sediment. Petrol. 23: 34–48.
Thomas, E. S. 1951. Distribution of Ohio animals. Ohio Jour. Sci. 51: 153–167.
Thomas, E. S. et al. 1952. The Orleton Farms mastodon. Ohio Jour. Sci. 52: 1–28.
Thorarinsson, S. 1943. Oscillations of the Iceland glaciers in the last 250 years. Geog. Analer. 25: 54 pp.
VerSteeg, K. 1933. The thickness of the glacial deposits in Ohio. Science. 78: 459.
Victor, P. E. 1953. Campagne au Groenland. Expeditions Polaires Francaise, Publ. Prelim. 16.
White, G. W. 1953. Sangamon soil and early Wisconsin loesses at Cleveland, Ohio. Amer. Jour. Sci. 251: 362–368.
White, S. E. 1956. Glaciological studies of two outlet glaciers, northwest Greenland, 1953. Medd. om Grønland. 137: 31.